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Complex yet translucent: the optical properties of sea ice

Donald K. Perovich*

Engineer Research and Development Center, Cold Regions Research and Engineering Laboratory, 72 Lyme Road, Hanover, NH 03755, USA

Abstract

Sea ice is a naturally occurring material with an intricate and highly variable structure consisting of ice platelets, brine pockets, air bubbles, and precipitated salt crystals. The optical properties of sea ice are directly dependent on this ice structure. Because sea ice is a material that exists at its salinity determined freezing point, its structure and optical properties are significantly affected by small changes in temperature. Understanding the interaction of sunlight with sea ice is important to a diverse array of scientific problems, including those in polar climatology. A key optical parameter for climatological studies is the albedo, the fraction of the incident sunlight that is reflected. The albedo of sea ice is quite sensitive to surface conditions. The presence of a snow cover enhances the albedo, while surface meltwater reduces the albedo. Radiative transfer in sea ice is a combination of absorption and scattering. Differences in the magnitude of sea ice optical properties are ascribable primarily to differences in scattering, while spectral variations are mainly a result of absorption. Physical changes that enhance scattering, such as the formation of air bubbles due to brine drainage, result in more light reflection and less transmission.
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Keywords: Sea ice; Optical properties; Albedo

1. Introduction

Sea ice is a naturally occurring material with an intricate and highly variable structure consisting of ice platelets, brine pockets, air bubbles, and precipitated salt crystals. A distinguishing feature of sea ice is that it is always at its melting point, because of the brine entrained during initial ice formation. Sea ice exhibits considerable spatial and temporal variability in its structure and its properties. Understanding the interaction of sunlight with sea ice is important to a diverse array of scientific problems, including those in climatology and biology. A key optical parameter for climatological studies is the albedo, the fraction of the incident sunlight that is reflected. Because of the positive feedback, changes in the albedo of sea ice may have a significant effect on the global climate [1]. The amount of photosynthetically active radiation (400–700 nm) and ultraviolet radiation (280–400 nm) transmitted through sea ice strongly affects biological activity in and under a sea ice cover [2,3]. This paper provides an overview of the material properties of sea ice, radiative transfer in sea ice, the ice albedo feedback mechanism, and the seasonal evolution of albedo.

*Tel.: 603-646-4255; fax: 603-646-4644.
E-mail address: perovich@crrel.usace.army.mil (D.K. Perovich).

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2. Sea ice

At their seasonal peaks, Arctic sea ice covers approximately 15,000,000 km^2 and Antarctic sea ice 20,000,000 km^2. The thickness of sea ice ranges from new ice only a few centimeters thick to pressure ridges tens of meters thick. The sea ice cover moves tens of kilometers in a day, driven by wind and currents. The ice forms a barrier, inhibiting the exchange of heat, salt, and momentum between the atmosphere and the ocean. For most of the year, sea ice is covered by a thermally insulating, highly reflecting snow cover.

On a small scale, sea ice is not a monolithic piece of pure ice. Rather, it is a structurally complex, and highly variable, mixture of ice, brine, air, and solid salts. For details regarding the physical and optical properties of sea ice, the reader is directed to several excellent review articles that discuss the initial growth of ice from the ocean [4] and the structural [5], physical [5,6], morphological [7,8], and optical [9] properties of the ice.

There are few key points that are particularly salient to the optical properties of sea ice. The salts in seawater do not become incorporated into the ice lattice. As a result, the ice grows as platelets (Fig. 1a) with pockets of brine trapped in between (Fig. 1c). The practical consequence of this is that sea ice is always at its melting point. Changes in ice temperature result in changes in brine volume by internal melting or freezing. If the ice warms, there is melting on the walls of the brine pockets until the salinity of the brine reaches equilibrium. Similarly, if the ice cools, there is freezing on the walls. The ice platelets are approximately 1 mm across. The brine pockets (Fig. 1c) are typically elongated ellipsoids roughly 0.1 mm across and tenths of a millimeter to millimeters long. The faster the growth rate, the smaller the platelets, and the more brine entrained in the ice. Air bubbles (Fig. 1b) tend to be spherical with a median diameter of approximately 1 mm [10]. If the ice is cold enough, solid salts precipitate, the two primary salts being mirabilite ($Na_2SO_4 \cdot 10H_2O$) at −8.2°C and hydrohalite ($NaCl_2 \cdot 12H_2O$) at −22.9°C.

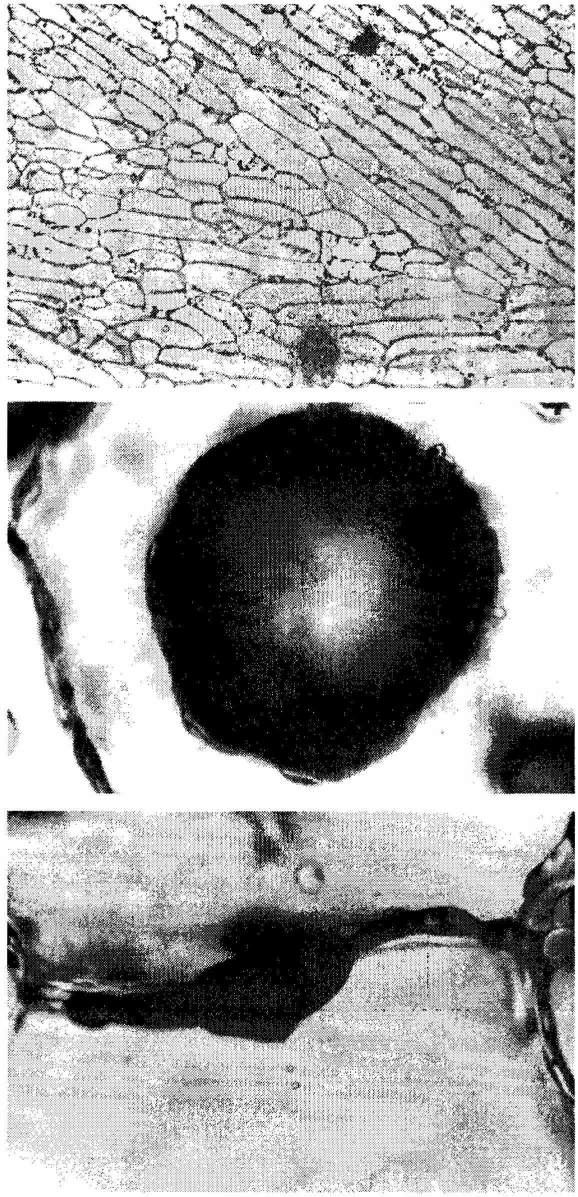

Fig. 1. A look at sea ice at three different scales focusing on (a) an array of ice platelets each approximately 1 mm across, (b) a single air bubble 1 mm across, and (c) a brine pocket 0.1 mm across.

The salinity of the brine and the precipitation of solid salts are governed by the equilibrium curve of the sea ice phase diagram [5]. There are empirical relationships between brine volume (v_b fraction), salinity (S‰), and temperature (T °C), such as the following [5]:

$$v_b = S\left(0.0532 - \frac{4.919}{T}\right). \quad (1)$$

As the ice warms and the brine volume increases, individual brine pockets connect, increasing the permeability and eventually allowing the brine to drain [11]. Sea ice floats with approximately 10% of its thickness above the surface of the ocean. As the brine drains from this portion of the ice, brine pockets transform into air bubbles, significantly affecting the optical properties of the ice. Changes in the small-scale structure of the ice, in the brine pockets, air bubbles, and solid salts, affect the scattering and absorption of sunlight in the ice.

3. Radiative transfer

At solar wavelengths, radiative transfer in sea ice is governed by absorption and scattering. In most treatments [12–15], the sea ice is generally considered to be a plane-parallel medium, where radiative transfer is defined by [16,17]

$$-\mu\frac{\mathrm{d}I(\tau,\mu,\phi,\lambda)}{\mathrm{d}\tau} = I(\tau,\mu,\phi,\lambda) - S(\tau,\mu,\phi,\lambda), \quad (2)$$

where I is the radiance, μ is the cosine of the zenith angle, and ϕ is the azimuth angle. τ is the non-dimensional optical depth and is defined as

$$\tau(\lambda) = (k(\lambda) + \sigma(\lambda))z,$$

where k is the absorption coefficient, σ is the scattering coefficient, and z is the physical depth. Scattering is included in the S term, which is expanded as

$$S(\tau,\mu,\phi,\lambda) = \frac{\varpi_0}{4\pi}\int_0^1\int_0^{2\pi} p(\mu,\mu',\phi,\phi',\lambda) \times I(\tau,\mu',\phi',\lambda)d\mu' d\phi - \frac{E_0(\lambda)}{4}p(\mu_0,\mu,\phi_0,\phi,\lambda)e^{-\tau(\lambda)/\mu_0}, \quad (3)$$

where $p(\mu,\mu',\varphi,\varphi')$ is the phase function, ϖ_0 is the single scattering albedo, and E_0 is the radiance of the direct beam component of the incident radiation field. The first term of Eq. (3) represents the scattering of the diffuse radiance field and the second term the scattering of the attenuated direct beam. The phase function represents the angular distribution of the scattered light.

There are three key inherent optical properties for sea ice: the absorption coefficient, the scattering coefficient, and the phase function. The absorption coefficient of sea ice is the linear sum of absorption coefficients of ice, brine, solid salts, and air weighted by their relative volume [13,18]. Since fresh ice is the primary component of sea ice, it dominates the absorption coefficient with a secondary contribution from brine. Absorption in the air is negligible and the contribution from solid salts is assumed to be relatively small. Absorption coefficients of pure ice [19,20] and clear Arctic water [21] are plotted in Fig. 2. Spectral changes in absorption coefficient are extremely large, spanning over three orders of magnitude from 470 to 1400 nm. Absorption coefficients of clean Arctic water are similar both in magnitude and in spectral shape to pure ice values. If particulates, sediments, ice biota, or dissolved organics are present in the

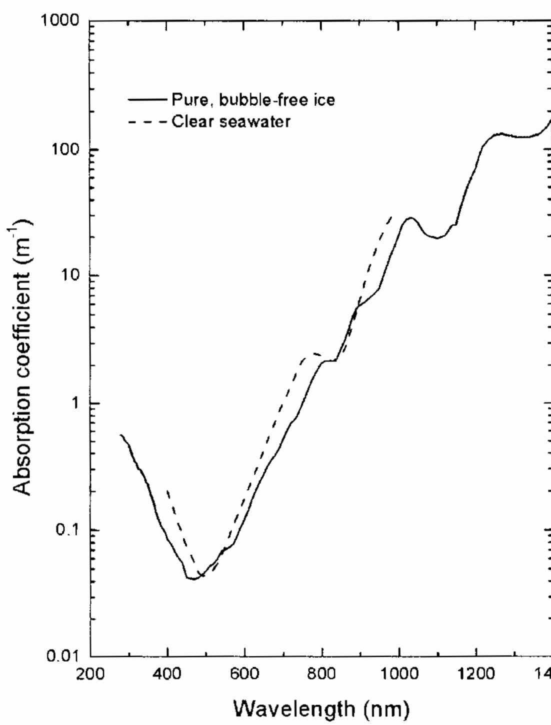

Fig. 2. Absorption coefficients of pure bubble-free ice [20,33] and clear Arctic water [21].

sea ice in sufficient quantity, then their absorptive properties must also be explicitly considered [3].

Sea ice has an abundance of air bubbles and brine pockets, and is a highly scattering medium. With a greater difference in index of refraction, air bubbles ($n \sim 1.0$ compared to $n \sim 1.31$ for ice) are more strongly scattering than brine pockets ($n \sim 1.34$–1.40 [22]). At lower temperatures, solid salts are formed [5] and scattering is significantly increased [23]. Since the air bubbles (~ 1 mm) and brine pockets (~ 0.1 mm) are much larger than the wavelength, contributions from diffraction and interference are ignored [12,24]. The wavelength dependence of the real portion of the index of refraction for ice, brine, and air is very weak at optical wavelengths. Thus, a major simplification is made by assuming that the scattering coefficients and phase functions are constant with wavelength [12,13]. Observational and theoretical studies indicate that scattering coefficients of sea ice are large, with values typically greater than $10\,\mathrm{m}^{-1}$ for warm ice and greater than $200\,\mathrm{m}^{-1}$ for bubbly ice or ice with precipitated hydrohalite [23]. Phase functions are strongly forward peaked [25,18]. Since there is a large amount of scattering in the sea ice, it is exceedingly difficult to directly measure scattering coefficients and phase functions. Values for these parameters are typically obtained from Mie scattering modeling of the inclusions, or by inference using a radiative transfer model to fit observations of multiply scattering samples.

4. Albedo

Examining the albedo (α) of sea ice provides a good illustration of the relationship between the physical properties of sea ice and radiative transfer. The albedo is the simplest, and in many ways the most significant, apparent optical property of sea ice. The albedo is the fraction of the incident irradiance that is reflected. For a perfectly white surface $\alpha = 1$, and for a perfectly black surface $\alpha = 0$.

The ice-albedo feedback mechanism plays a key role in the surface heat budget of Arctic sea ice [26,27]. Furthermore, understanding the ice-albedo feedback is necessary to accurately assess the role of the Arctic sea ice cover in global climate change [1,28–30]. The ice-albedo feedback begins with snow-covered sea ice, one of the most highly reflecting materials on the earth. However, some of the incident solar energy is absorbed, first warming, then melting the surface. As melting progresses, the snow-covered ice is transformed into a darker, lower albedo, combination of open ocean, melt ponds, and bare ice, allowing more solar energy to be absorbed and melting to accelerate. This is a positive feedback and is of interest to studies involving climate change because of its capacity to amplify small changes.

Because of the importance of albedo, there is a large, comprehensive database of spectral and wavelength-integrated sea ice albedos [9,31–36]. From a climatic perspective, it is important not just to have a catalog of albedos, but also to understand how the large-scale albedo (α_L) of the Arctic basin evolves over the annual cycle. Ideally, the large-scale albedo would be measured directly by satellites. Unfortunately, the utility of satellite measurements in the summer is severely limited by the pervasive cloud cover and by the widespread presence of surface water. Another method of estimating albedo is to assume that the surface is composed of several surface types (S_i). Each of these surfaces has an albedo (α_{Si}) and covers a fraction of the total area (A_{Si}). The large-scale albedo is simply the sum of the individual surface-type albedos weighted by their areal coverage, or

$$\alpha_L = \sum_i \alpha_{Si} A_{Si}. \tag{4}$$

Estimates determined in this manner are a combination of surface observation, satellite measurements, model results, inference, and intuition.

For much of the year, estimating the large-scale albedo (α_L) is relatively straightforward. From November to March, the albedo is of little importance, since it is the long Arctic night and the incident irradiance is small or zero. By March, there is an appreciable amount of sunlight, and by April, there are 24 h of daylight at high latitudes. From March through late May, the surface is primarily sea ice covered by an optically thick layer of snow, with a small fraction of the surface consisting of the open ocean. Since the albedos of snow and the open ocean are well known, all that is needed to compute the large-scale albedo is a fraction of the open water. This fraction is routinely determined from satellite data. With the onset of melt in late May to early June, the situation becomes considerably more complex, as the surface becomes a variegated mix of melting snow, bare ice, ponded ice, and melt ponds. The melting snow is ephemeral, melting away in just a couple of weeks. For much of the summer, the primary surface types are bare ice, ponds, and open ocean. Pegau and Paulson [37] established that the open ocean albedo was about 0.07 and temporally invariant throughout the summer. Photographs showing seasonal changes of the sea ice surface (a) in April before melt, (b) in June early in the melt season, and (c) in August near the end of melt are presented in Fig. 3.

The evolutionary sequences of bare ice and a melt pond measured from April to August 1998 are presented in Fig. 4. These observations were made as part of a year-long field experiment examining the surface heat budget of the Arctic Ocean [38]. Since both the sites were covered by an optically thick snow layer, the albedo was the same until melt began in late May. After only a few days of melting, the albedos diverged, as the pond site was first to melt free of snow. There was an occasional increase in albedo (e.g., 26 June), during brief periods of subfreezing temperatures and new snow.

The bare ice albedo is remarkable for its constancy, maintaining a value of 0.65 ± 0.05 over the course of the summer. Examining the bare ice portion of Figs. 3(b) and 3(c), the surface appears to be unchanged, even though there was 0.5 m of surface ablation. The stars in Fig. 4 denote the albedo when Fig. 3b (0.64) and Fig. 3c (0.67) were taken. The remarkably constant bare ice albedo was attributable to a regenerating, few centimeter thick surface scattering layer of granular, decomposing ice. As sea ice is translucent, a portion of the incident sunlight penetrates the ice. The penetrating sunlight is absorbed preferentially at the grain boundaries, brine pockets, and air bubbles and causes melting. Because the ice is permeable at this time, the meltwater drains and the ice fragments into small pieces. These small pieces form the surface scattering layer and are similar in appearance to coarse-grained snow [39,40]. As the melt season progressed, this layer

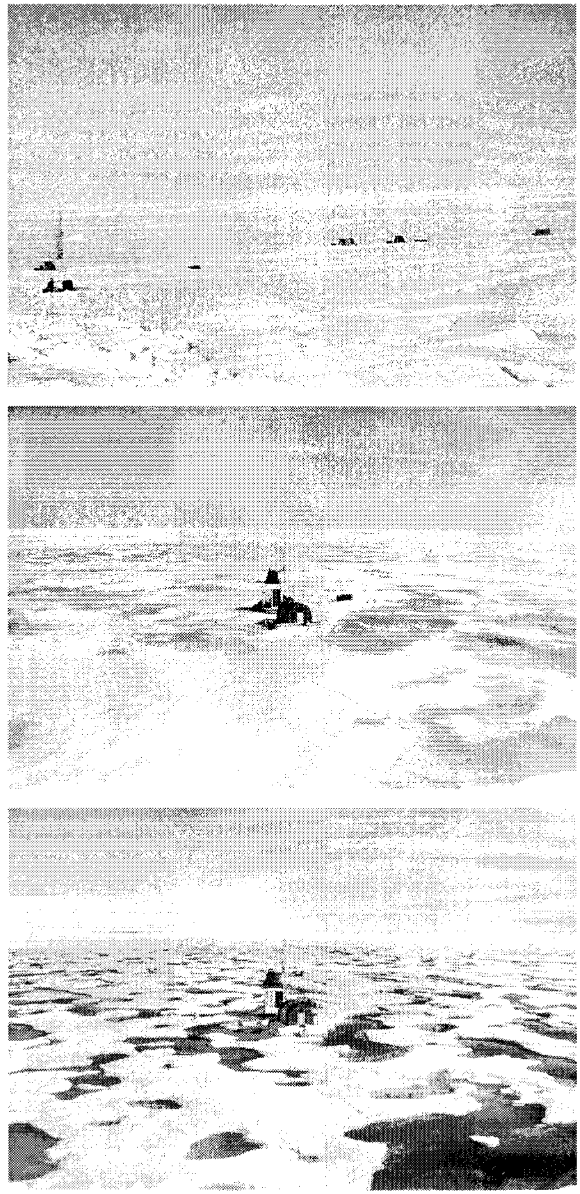

Fig. 3. Evolution of ice surface from spring through summer melt. The photographs were taken (a) well before the onset of melt in mid-April 1998, (b) early in the melt season on 17 June 1998, and (c) near the end of the melt season on 7 August 1998.

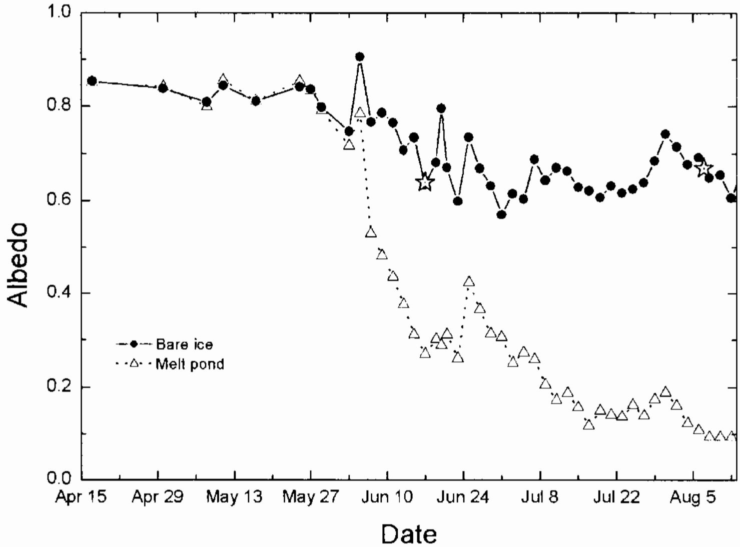

Fig. 4. Seasonal evolution of albedo for bare ice and ponded ice. The stars denote bare ice albedos measured when photographs 3b and c were taken.

kept renewing itself because of melting within the near-surface portion of the ice, resulting from the ongoing radiation absorption in the upper layers of the ice [40]. The thickness of the surface scattering layer ranged from 1 to 3 cm. Foggy days with condensation melting on the surface, tended to thin the scattering layer, and sunny days with larger fluxes of penetrating solar irradiance, tended to increase the layer thickness [36]. The constant albedo of bare ice is a key feature and simplifies estimation of large-scale albedos for the ice pack.

Unlike bare ice, the melt pond albedo exhibited a strong temporal dependence. With the exception of the brief cooling events, the melt pond albedo decreased throughout the summer. It reached a minimum value of 0.1 in mid-August just before freezeup began. Physical changes in the ponds are evident in Figs. 3(b) and 3(c) as the ponds on 7 August appear better defined and much darker. Pond physical property measurements indicated that there was approximately 50% more surface melting on ponded ice compared to bare ice, and that the pond water depth increased throughout the summer. The darker appearance and the lower albedo were not due to the increase in water depth. The pond water was clear and increasing the depth from 5 to 40 cm made little difference to the albedo. What made a difference was the continued melting of the ice underneath the surface water. As this ice thinned, there was less backscattering and the albedo decreased. The albedo of the ponds is primarily ascribable to the optical properties of the underlying ice, not the pond depth [36]. Understanding pond evolution, how ponds deepen and how they change their areal extent, is critical to determining the large-scale albedo. The physics underlying pond evolution is complex and just beginning to be understood [41]. Factors affecting pond evolution include albedo, snow depth, surface melt rate, topography, ice structure and ice permeability.

5. Summary

Sea ice is a structurally complex, highly scattering medium that is always at its melting point. The interaction of solar radiation with sea ice is directly related to changes in the state and structure of the ice. Physical changes in the ice, which enhance scattering, such as the formation of air bubbles because of brine drainage, result in larger albedos and less light is transmitted. Radiative transfer in sea ice is a combination of absorption and scattering. Differences in the magnitude of the optical properties of sea ice are primarily due to differences in scattering, while spectral variations are mainly the result of absorption. More theoretical and observational studies are needed to improve our understanding of scattering in sea ice.

Sea ice albedo is extremely sensitive to the surface state. If the ice has an optically thick (5–10 cm) snow cover, wavelength-integrated albedos are about 0.85. The albedo decreases as the snow melts away. The presence of liquid water on the surface causes a decrease in albedo, which is more pronounced at longer wavelengths. If the ice surface drains, the brine pockets become air bubbles, resulting in more scattering and an increase in albedo. The albedo of bare, multi-year ice changes little during the melt season, because of the continued presence of a highly scattering surface layer. In contrast, melt pond albedos decrease throughout the summer as the ponds grow deeper and the underlying ice thins. The greatest uncertainty in modeling the evolution of sea ice albedo is understanding how the ice albedo, melt rate, surface topography, and ice permeability affect melt pond formation and development.

Acknowledgements

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